37p

CONTACT METAMORPHISM IN THE

MT. TALLAC ROOF REMNANT

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INTRODUCTION

A large remnant of the roof of the composite Sierra Nevada

batholith is exposed at the crest of the range within the Fallen Leaf

Lake 15' quadrangle (Figure 1). The roof remnant consists of a

15,000-foot section of Triassic(?) and Jurassic metasedimentary and

metavolcanic rocks which have been intruded by a dozen stocks and

batholiths and numerous smaller bodies. The remnant is two to four

miles wide and about 10 miles long. The plutons range in composition

from gabbro to granite, although granodiorite bodies are volumetrically

by far the most important.

A 7600-foot sequence of meta-andesite and metabasalt tuffs,

tuff-breccias, and flows conformably overlies a 7500-foot sequence of

marine and fluvial sandstones, siltstones, and conglomerates within

the remnant. Carbonate rocks and highly aluminous shales are

conspicuously absent from the sequence. Figure 2 is a geologic map

showing the gross distribution and structural trends of the metamorphic

rocks and the distribution and composition of the plutons.

<u>Facies designations</u>. — The distribution of the metamorphic mineral facies within the roof remnant are plotted in Figure 2. Contact metamorphic mineral facies are keyed on mafic assemblages only. The facies designations used are those of Turner (Fyfe, Turner, and Verhoogen, 1958) except for the substitution here of the actinolite hornfels facies for the more common albite-epidote hornfels facies. Mafic rocks containing albite plus epidote are not as common in the Mt. Tallac remnant as those containing actinolite, greenish biotite, and more calcic plagioclase.

As used in this paper, a metamorphic facies comprises all of the rocks coexisting with mafic rocks which contain a given key mineral assemblage. Changes in key mineral assemblages are plotted on the map as changes in mineral facies. These mineral assemblages developed under some range of temperature, total pressure, and vapor phase composition, and for rocks of mafic composition within any given facies, the range of P, T conditions was probably closely similar throughout the area.

Important changes in mineral assemblages in rocks of other compositions may or may not coincide in the field with the changes in mafic rocks. For example, the calcareous assemblage <u>grossularite-quartz</u> (<u>-calcite</u>) is found in the actinolite-hornfels facies in the Mt. Tallac rocks but in the hornblende-hornfels facies elsewhere. In the Mt. Tallac remnant, the start of the reaction to produce grossularitic garnet in calcareous rocks occurred farther from the contacts than the start of the actinolite to hornblende reaction in mafic rocks. Grossularite is commonly considered to be correlative with the hornblende hornfels facies

in most other areas. Also, the assemblage diopside-quartz-calcite

(-actinolite-clinozoisite) occurs with actinolite-bearing mafic rocks

in the Mt. Tallac remnant; the appearance of diopside is not correlative

with the low-grade limit of the hornblende hornfels facies as determined

by mafic rocks.

ROCKS FARTHEST FROM THE CONTACTS

The lowest grade rocks within the roof remnant have all been

more or less recrystallized. The fine-grained matrices of both

metavolcanic and metasedimentary rocks have recrystallized to minerals

distinctive of the albite-epidote hornfels facies except for the

widespread metastability of calcic plagioclase in place of the pair

albite-epidote. Chlorite is much less common than in typical

greenschist terrains. Pumpellyite has not been recognized.

Most rocks have not been deformed internally and are not

schistose except locally near some igneous contacts and in some fault

zones.

Mafic and intermediate rocks. -- Silica-saturated mafic and

intermediate assemblages farthest from the contacts are

plagioclase (An_{5-60})-actinolite-quartz-green to tan biotite-ore.

Much less common is the assemblage

albite-actinolite-epidote-biotite(chlorite)

which is found in most areas elsewhere. Silica-deficient rocks commonly

contain a tan amphibole with refractive indices less than 1.66 and high $2V_x$.

Relict clinopyroxene is present in some uralite pseudomorphs (Figure 3).

The breakdown of clinopyroxene to actinolite (largely a volume for volume replacement) occurred generally without the usual breakdown of plagioclase to epidote and albite. For this reason, the low-grade rocks have been assigned to the actinolite hornfels facies.

Plagioclase is rarely albite; it is clear to turbid, and the very fine-grained clouding is concentrated in the calcic zones. The plagioclase coexisting in the actinolite rocks is not recrystallized; original phenocrysts retain their euhedral shape and oscillatory zoning. A possible explanation for the continued metastability of calcic plagioclase is that excess sodium that would normally have been derived from sea water was absent in this sequence of non-marine rocks. Commonly-observed relations in greenstones elsewhere indicate that albite replaced calcic plagioclase volume for volume. Therefore, sodium must have been derived from source outside the plagioclase itself because the mere breakdown of calcic plagioclase to albite plus epidote is a reaction with a decrease of volume of nearly 20 percent.

<u>Calcareous metasedimentary rocks</u>. -- Assemblages coexisting

with actinolite-bearing mafic rocks are, with increasing CaO:

a) <u>quartz-actinolite-albite-biotite</u>

b) <u>quartz-actinolite-diopsidic clinopyroxene-albite-</u>

    <u>(biotite-epidote)</u>

c) <u>quartz-diopside-albite-(calcite-epidote)</u>.

These assemblages occur interbedded with one another, and therefore

the control on the presence of an amphibole or a clinopyroxene is

very sensitive to composition. Figure 4 shows coexisting diopsidic

augite and a pale green actinolitic(?) amphibole. Other important

minerals are quartz, clinozoisite, and calcite. The relation of

clinopyroxene to CaO activity is discussed below.

Some metasedimentary rocks contain patches of albite up to 0.05 mm across which irregularly replace detrital plagioclase; it is not known at what time in the history of the rocks the albite was introduced. Epidote-replaced clasts in the conglomerate and epidote nodules locally present in the tuff-breccias make epidote a volumetrically important mineral. Epidote occurs with or without magnetite and quartz in veins, nodules, and along shear zones in metavolcanic rocks near igneous contacts. The epidotization is clearly related to the intrusive rocks. The best area is that adjoining the northern contact of the Keiths Dome quartz monzonite from Fallen Leaf Lake to Glen Alpine. Veins of quartz, epidote, magnetite, and tourmaline are common. The epidote was introduced into the wallrocks early in the history of the intrusion of any particular granitic body because the nodules and veins are themselves contact metamorphosed.

HORNBLENDE HORNFELS FACIES ROCKS

Delimiting the Facies

The line delimiting the low-grade boundary of the hornblende-hornfels facies as here described is defined by the change of actinolite to hornblende in mafic rocks. The plagioclase composition is not an indicator because it was not albite prior to recrystallization. The optical and chemical limits of actinolite in rocks of normal basaltic compositions have not been defined. In an effort to be consistent in decisions regarding recognition of the amphiboles, three different criteria were used in petrographic determinations:

a) fibrous unoriented pale green masses of "uralite" that have relict cores of pyroxene are called actinolite;

b) optically oriented recrystallized pseudomorphs are called actinolite if they have multiple twinning on (100);

c) amphiboles that are pale green, blue-green, or yellow-green in Z are called actinolite if they have n_Z less than 1.660.

The assemblage hornblende-plagioclase changes to hornblende-diopsidic augite-plagioclase in the innermost zones of the aureoles at Glen Alpine and Dicks Lake. Accordingly, the line marking the first appearance of metamorphic clinopyroxene in mafic rocks has been plotted in Figure 2 where control exists.

Mineral Assemblages and Textures

Mafic and intermediate rocks. -- Undersaturated rocks have the assemblage labradorite-greenish tan hornblende-red brown biotite-ore. Saturated mafic rocks have the first of the following assemblages almost universally; they develop the last two assemblages only locally and at higher grades:

An$_{30-50}$-green hornblende-greenish brown biotite-quartz(-ore)

An$_{30-50}$-green hornblende-brown biotite-diopsidic augite-quartz

An$_{30-50}$-brown biotite-diopsidic augite-quartz.

It is important to recognize the difference between prograde and retrograde hornblende. Figure 5 shows the clean, unsieved texture of prograde hornblende; Figure 6 shows complexly poikiloblastic retrograde hornblende which has developed after diopsidic augite. Relict plagioclase textures are retained even in the highest grade rocks at Glen Alpine Springs (Figure 7).

The metamorphic augite that is farthest from the contact in the mafic rocks occurs in isolated grains in small replacement veins of quartz, K feldspar, and medium oligoclase. The first appearance of the augite is about 200 to 250 feet from the contact at Glen Alpine Springs. It is locally distributed around the contact of the pluton at Dicks Lake. All of the metavolcanic rocks around the granodiorite at Glen Alpine Springs have developed augite but none contain hypersthene, even in the most mafic lapilli in the tuff-breccias. These relations indicate that the presence of augite rather than hypersthene at this contact is due to insufficient temperature, not merely to a high CaO content.
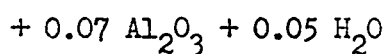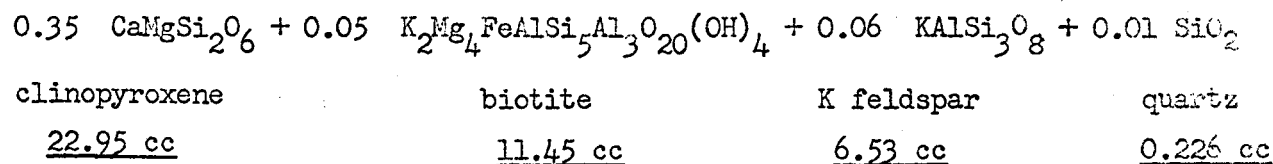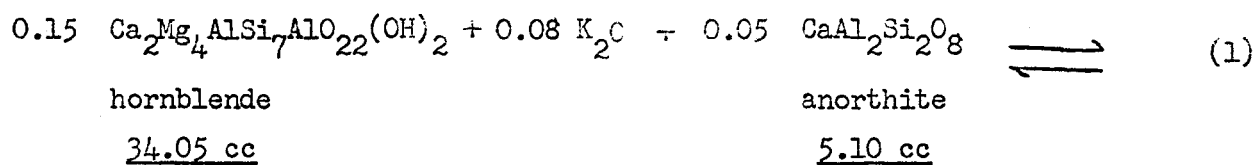
Table 1 contains modes of five typical metabasalts collected at different distances from the east contact of Glen Alpine granodiorite. Biotite, K feldspar, and augite increase toward the contact at the expense of hornblende and anorthite. The changes in quartz and ore (magnetite) are sporadic and dependent on both original composition and retrograde oxidation of specimen B. The actual mineral changes in cc from the hornblende-biotite specimen 444 to the augite-biotite specimen B are listed in the last column of the table. Those two samples are used for the comparison because they match most closely in chemical composition. The change in volume of plagioclase from 50 to 43 percent between 444 and B indicates that plagioclase entered the reaction to form the augite. The following simplified reaction approximates the complex total reaction which took place during the formation of the augite-biotite hornfelses from the hornblende-biotite hornfelses:

Table 1. Modes of five metavolcanic hornfelses at different distances from the
Glen Alpine granodiorite, and analyses of comparison specimens 444 and B.

Distance from nearest contact	1800 feet	400 feet	350 feet	200 feet	30 feet	modal change:
Specimen no.	94	444	438	439	B	B − 444
quartz	1	6.5	4.5	6.4	1.2	− 5.3
plagioclase	50	50.0	50.0	51.7	43.3	− 6.7
K feldspar				0.1	4.8	+ 4.8
hornblende		34.0	27.6	19.2	1.1	− 32.9
biotite	3	9.0	16.4	20.7	19.4	+ 10.4
diopsidic augite			0.1	0.1	25.8	+ 25.8
actinolite	42					
ore	3	0.5		1.2	4.2	
epidote	1					
sphene	tr		1.2	0.5	0.2	
total	100	100.0	99.8	99.9	100.0	

SiO_2		54.41			54.05
Al_2O_3		18.29			16.79
TiO_2		0.93			0.71
Fe_2O_3		1.70			3.43
FeO		7.32			6.24
MgO		3.92			4.61
CaO		8.14			7.09
Na_2O		3.13			3.09
K_2O		1.40			3.20
H_2O^+		nd			nd
total		99.24			99.21

Plagioclase composition	An_{40}	An_{45}	An_{50}	An_{40}	An_{40}
Density	2.935	2.895	2.855	2.850	2.870

Analyses by A. G. Loomis, Berkeley, California

$$0.15 \ Ca_2Mg_4AlSi_7AlO_{22}(OH)_2 + 0.08 \ K_2O + 0.05 \ CaAl_2Si_2O_8 \ \rightleftharpoons \quad (1)$$

hornblende   anorthite

<u>34.05 cc</u>   <u>5.10 cc</u>

$$0.35 \ CaMgSi_2O_6 + 0.05 \ K_2Mg_4FeAlSi_5Al_3O_{20}(OH)_4 + 0.06 \ KAlSi_3O_8 + 0.01 \ SiO_2$$

clinopyroxene   biotite   K feldspar   quartz

<u>22.95 cc</u>   <u>11.45 cc</u>   <u>6.53 cc</u>   <u>0.226 cc</u>

$$+ \ 0.07 \ Al_2O_3 + 0.05 \ H_2O$$

water

<u>1.84 cc</u>

Total volumes at 600°C and 1400 bars: 39.15 cc → 43.00 cc

Table 2 compares the solid volumes involved in the reaction which were computed from the modal data with those in the reaction as written above. The correspondence is reasonably good and could be made to be closer by artitrarily adjusting the compositions of the hornblende and biotite and adding iron to the model equation. The important point is that the zone of K_2O introduction coincides closely with the appearance of diopsidic augite. The reaction of hornblende to augite in rocks of any given CaO content apparently may take place at lower temperatures when K feldspar and biotite are also possible products.

Table 2. Comparison of modal and calculated volumes involved in hornblende to augite reaction shown by specimens 444 and B.

	hornblende	+ quartz	+ anorthite	→
modal	32.9	5.3	6.7	
calculated	34	~ 0	5.1	

	augite	+ biotite	+ K feldspar
modal	25.8	10.4	4.8
calculated	23	11.5	6.5

Calcareous rocks. -- The assemblage

quartz-diopside-grossularite-plagioclase is commonly considered to have

formed under the same conditions as hornblende-bearing mafic hornfelses.

Rocks with grossularitic garnet coexist with actinolite-bearing mafic

rocks in the Mt. Tallac roof remnant, however. The best examples of

such rocks are specimens 364, a tuff-breccia, and 368, a metaconglomerate.

They occurred about 300 feet apart and the tuff-breccia was 150 feet

closer to the contact of the Dicks Lake pluton. The metaconglomerate is

much more recrystallized than the tuff-breccia; it is a heterogeneous

rock that contains about 40 percent quartz, 20 percent plagioclase, and

nearly equal amounts of amphiboles, diopsidic augite, calcite, and

grossularitic garnet, with minor epidote and sphene. The garnet is in

masses with calcite and clinopyroxene that form the matrix of the pebbles.

The tuff-breccia specimen contains no hornblende, yet it was

150 feet closer to the contact than the metaconglomerate. This relation

indicates one or more of the following.

a) The rocks are equilibrium assemblages of the albite-epidote-
hornfels facies.

b) A meaningful facies designation cannot be given to the meta-
volcanic rock because the common low-grade aluminous minerals
chlorite and epidote were not present to facilitate the
reaction of actinolite to hornblende.

c) The temperature could have been higher at a given distance from the contact in the metaconglomerate unit than in the metavolcanic unit for two reasons. First, the thermal diffusivity of the former is about twice that of the latter. Using average values for thermal conductivity k for the (quartzitic) mataconglomerate and the tuff-breccia of 10×10^{-3} and 6×10^{-3} $\frac{cal}{sec\ cm\ deg}$ respectively (Birch, et al., 1942, p. 253-256), densities ρ of 2.75 and 2.90, and a heat capacity c of 0.22 for both, the diffusivities ($\alpha = \frac{k}{\rho c}$) are 16.5×10^{-3} for the metaconglomerate, and 9.4×10^{-3} for the tuff-breccia. Second, the conglomerate unit must have been much more permeable to heated water being forced through the walls away from the hot pluton. Much more water per unit time would be expected to travel through the metaconglomerate; the water would lose heat to the containing rocks by condensation.

d) f_{H_2O} was so high relative to f_{CO_2} that the actinolite to hornblende reaction was inhibited while the calcite plus anorthite or epidote were able to begin reacting to garnet.

One occurrence of wollastonite is in the aureole of the gabbro at Lake Schmidell. Assemblages there are <u>calcite-wollastonite-grossularite-bytownite(-vesuvianite)</u>. These rocks contained graphite originally, but it is not pre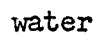sent in the high-grade hornfelses. It is possible that the carbon was oxidized at the contact temperature by hydrogen from water, but the resultant increase in f_{CO_2} would only tend to inhibit the wollastonite reaction. Wollastonite also occurs in the calcareous metaconglomerate east of Dicks Peak out to about 300 feet from the contact. Mafic rocks are associated with wollastonite hornfelses southwest of Dicks Lake where the former are hornblende-hornfels facies rocks. In association with both hornblende-hornfels and pyroxene-hornfels facies mafic rocks elsewhere in the area, rocks calcareous enough for wollastonite are absent, and the assemblages there are combinations of any or all of the following minerals quartz, diopside, grossularitic garnet, plagioclase An_{40-90}, K feldspar, scapolite, and sphene.

Reaction of epidote to andradite. -- Epidote nodules 2 to 10 inches
in diameter in the tuff-breccias have partially recrystallized to
andradite-rich garnet in the hornblende-hornfels facies rocks at
Glen Alpine and in the pyroxene-hornfels facies rocks at Dicks Lake.
The nodules appear to be concretionary replacement features; others
retain relict bedding and were epidotized sandstone clasts. The nodules
have reacted with the enclosing metavolcanic rocks in the aureoles,
adding CaO to them and thereby promoting the change from hornblende to
clinopyroxene. The following assemblages are found in and around a
nodule (specimen 741) from a mafic andesite tuff-breccia from the
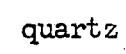wallrocks of the Glen Alpine granodiorite:

center of nodule	epidote-quartz
↓	andradite-anorthite-quartz (minor hematite)
rim of nodule	bytownite-magnetite-quartz
reaction jacket in tuff-breccia	labradorite-diopsidic augite-quartz (Some other nodules have hedenbergite)
tuff-breccia	andesine-hornblende-diopsidic augite-biotite-quartz.

The epidote contains 13.31% Fe_2O_3, equivalent to about 25% ferric end
member. The epidote from several other nodules is quite similar, on the
basis of refractive index. The garnet has $a_O = 12.034$ and $n_D = 1.87$,
corresponding to about $Andradite_{90}Grossularite_{10}$ (Skinner, 1956; ASTM Data
Cards).

The reaction of epidote to andradite plus anorthite is not strictly stoichiometric; it requires quartz and CaO or removal of Al_2O_3. Two of the alternative reactions are:

$$8\ Ca_2(Al_{0.75}Fe^{3+}_{0.25})_3Si_3O_{12}(OH) + 3\ SiO_2 + 2\ CaO \rightleftharpoons \tag{2}$$

$\qquad$ epidote $\qquad\qquad\qquad\qquad$ quartz

$$9\ CaAl_2Si_2O_8 + 3\ Ca_3Fe^{3+}_2Si_3O_{12} + 4\ H_2O$$

$\quad$ anorthite $\qquad$ andradite $\qquad\quad$ water

and

$$8\ Ca_2(Al_{0.75}Fe^{3+}_{0.25})_3Si_3O_{12}(OH) \rightleftharpoons 7\ CaAl_2Si_2O_8 + 3\ Ca_3Fe^{3+}_2Si_3O_{12} \tag{3}$$

$\qquad\qquad\qquad\qquad\qquad\qquad$ anorthite $\qquad\qquad$ andradite

$$+\ 2\ Al_2O_3 + SiO_2 + 4\ H_2O$$

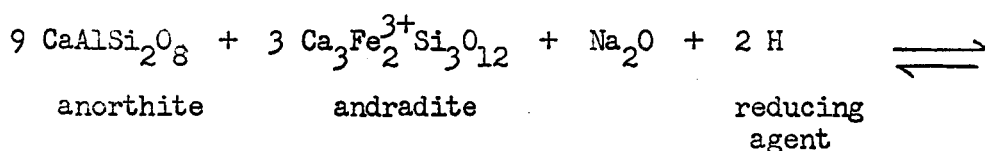$\qquad\qquad\qquad\qquad\qquad\qquad\qquad$ quartz

The iron-aluminum balance in the andradite of equation (3) could be adjusted so that a minor amount of hematite appears as a product, as is the case in the actual metamorphic assemblage. The reaction at the initial stage of recrystallization was probably close to reaction (3).
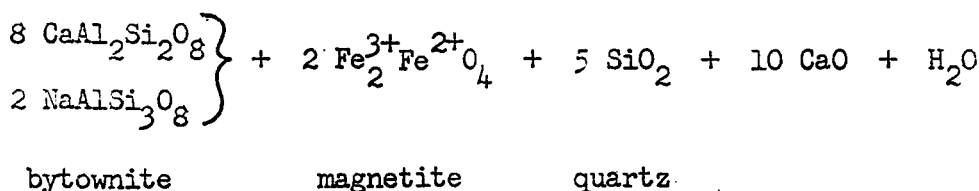
Some preliminary attempts to reproduce the reaction at 2000 bars P_{H_2O} have been made, using charges in platinum capsules surrounded by hematite as a hydrogen sink. (See Eugster and Wones, 1962, p. 88.) Starting materials have included (a) the interiors of natural epidote nodules, (b) natural epidote of appropriate composition with excess quartz, (c) synthetic oxide mixtures, and (d) glasses made from oxide mixtures. Runs from three weeks to six months have been made. An equilibrium between 575° and 600°C appears to have been established for a reaction much like reaction (3) above except that hematite rather than corundum appears in the products. The garnet is consequently more aluminous than the natural andradite; a_0 is commonly within the range 11.97_5 to 12.00_5, corresponding to Andradite$_{60-75}$ Grossularite$_{40-25}$. Runs above 600°C at 2000 bars P_{H_2O} produce golden-brown garnet euhedra, anorthite, and hematite. One successful run at 575°C produced yellow-green epidote euhedra but neither garnet, anorthite, nor hematite.

The assemblage <u>bytownite-magnetite-quartz</u> at the rim of the nodule shows an exchange with the enclosing rock in which some CaO went to the tuff-breccia and some Na_2O went to the nodule:

$$9\ CaAlSi_2O_8\ +\ 3\ Ca_3Fe_2^{3+}Si_3O_{12}\ +\ Na_2O\ +\ 2\ H \rightleftharpoons$$

anorthite   andradite   reducing agent

(4)

$$\left.\begin{array}{l} 8\ CaAl_2Si_2O_8 \\ 2\ NaAlSi_3O_8 \end{array}\right\}\ +\ 2\ Fe_2^{3+}Fe^{2+}O_4\ +\ 5\ SiO_2\ +\ 10\ CaO\ +\ H_2O$$

bytownite   magnetite   quartz

The excess CaO may have been used both to promote the breakdown of epidote by reaction (2) and to promote the reaction of hornblende to augite in the surrounding tuff-breccia. (See discussion on CaO control of clinopyroxene appearance later in the paper.)

<div align="center">PYROXENE HORNFELS FACIES ROCKS</div>

Pyroxene hornfels facies rocks were mapped at the appearance of hypersthene in mafic rocks. The hornblende-hornfels facies rocks around the quartz diorite at Dicks Lake change into the higher-grade hornfelses where the rocks of the roof remnant previously had been dried by the Camper Flat granodiorite. The compositional range of the high-grade rocks is small; the assemblages from the most mafic to the most calcic are:

a) labradorite-tan hornblende-hypersthene- red brown biotite (undersaturated)

b) labradorite-hypersthene- red brown biotite-quartz- K feldspar

c) labradorite-hypersthene-diopsidic augite- red brown biotite-quartz- K feldspar

d) andesine-diopsidic augite- red brown biotite-quartz- K feldspar

e) quartz-plagioclase An_{60-90}-diopside-garnet- K feldspar-scapolite.

None of the rocks contains ore.

The biotite-hypersthene reaction. -- An interpretation of the

entire paragenesis will be made by examining one rock closely. The

biotite-hypersthene reaction appears to be simple because it relates to

only one pyroxene and one hydrous silicate. The reaction relation of

biotite to hypersthene is not evident except under low magnification,

as shown by a comparison of Figures 8 and 9. Biotite and hypersthene

crystallized together, as is indicated by the clean crystal borders and

size of the biotite tablets, and the lack of a direct reaction relation

wherein hypersthene might jacket or otherwise obviously crystallize

directly from biotite. The rock was undergoing continuous recrystalliza-

tion as the following reaction proceeded slowly to the right:

$$K_2Mg_3Fe_3Si_6Al_2O_{20}(OH)_4 \ + \ 6 \ SiO_2 \ \rightleftharpoons$$

$$\text{biotite} \qquad\qquad\qquad \text{quartz} \tag{5}$$

$$3 \ MgFeSi_2O_6 \ + \ 2 \ KalSi_3O_8 \ + \ 2 \ H_2O.$$

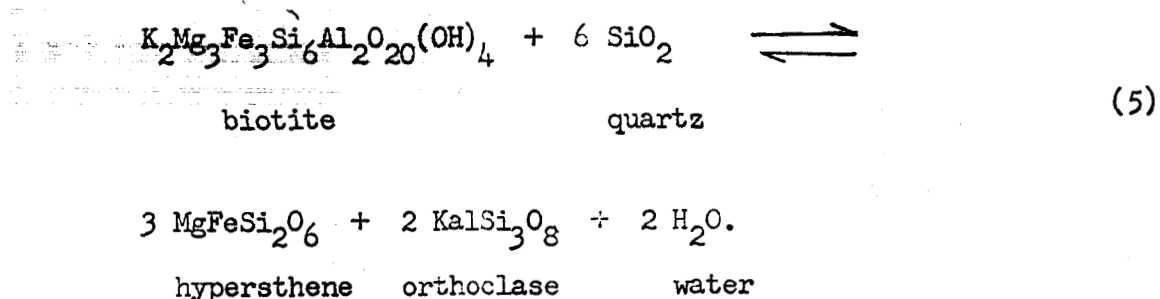$$\text{hypersthene} \quad \text{orthoclase} \qquad \text{water}$$

Table 3 lists the mode and analysis of the specimen for which

the reaction above was written and a calculated analysis of the biotite

in the rock. Although the mode shows no K feldspar there is abundant

K feldspar in immediately adjacent rocks; apparently it was very mobile

during recrystallization and tended to concentrate (Figure 10). The

biotite has $n_Z = 1.649 \pm 0.003$, birefringence 0.050 from thin section,

Z = red-brown, $2V_x$ less than 10°, ρ about 3.08. The red-brown color and

high TiO_2 are common in high-grade metamorphic biotites (eg. Engel and

Engel, 1960, p. 24).

Table 3. Mode and analysis of pyroxene hornfels specimen 613 and calculated analysis for the biotite.

	volume percent			weight percent	
				rock	biotite
quartz	2.3		SiO_2	52.2	36.7
plagioclase	59.0		Al_2O_3	19.6	16.6
K feldspar	0.0		TiO_2	0.8	4.4
biotite	20.2		Fe_2O_3	0.27	1.3
hypersthene	18.5		FeO	8.59	16.0
	100.0		MnO	nd	
plagioclase An_{56}			MgO	7.4	11.7
hypersthene En_{56}			CaO	6.0	0.7
			Na_2O	2.8	
Density 2.890			K_2O	1.9	8.6
			H_2O^-	nd	
				99.5	96.0

A = 31.8

C = 17.8

F = 50.4

Analysis by A. G. Loomis and the writer.

The equilibrium of the reaction above is dependent upon the partial pressure of water, which appears as a free phase, as well as temperature. Water pressure will inhibit the reaction at any given temperature when it rises to the point at which hypersthene is just stable with respect to a hydrous phase. Boyd (1959, p. 389) believes that subsolidus pyroxene is not stable at any temperature at a water pressure of more than about 1400 bars. A pressure of nearly 1400 bars was the approximate load pressure calculated from other considerations (see Physical Conditions). The reaction as written runs, at 720°C and 1400 bars, with an increase of volume from $435 \frac{cc}{mol} \longrightarrow 512 \frac{cc}{mol}$; the water at that temperature and pressure has a volume of $48.4 \frac{cc}{mol}$ (Holser and Kennedy, 1959, p. 76). For the reaction to run at equal or decreasing volume, about 80 percent of the water produced in the reaction must escape. Unless the reaction runs to a decrease of volume, the water pressure must equal the load pressure because the water vapor must hold the grains apart. If the reaction had been able to go to the right when water pressure was as great as load pressure, it

would have run until one of the reactants had been depleted; the only

control on the rate would have been the kinetics of the reaction.

It is concluded that the reaction could not run when water pressure was

as great as load pressure because the two reactants, biotite and quartz,

are still present in the rock and the textures in Figure 9 suggest

much recrystallization. The rate of the reaction must have been dependent

upon the rate at which water vapor flowed from the terrain. Yoder (1955,

p. 513-516) described the process as depending primarily upon the

permeability of the recrystallizing rock, provided that the reaction rate

is faster than the flow rate.

The problem of total vapor pressure in rocks undergoing recrystalli-

zation is very complex. It should be independent of volume under solid

state recrystallization if the load pressure at that depth far exceeds

the crushing strength of the mineral grains of the rock--that is, if the

rocks are not strong enough to support pore space, any gas phase must

support the rocks to exist. Surface phenomena or the presence of anatectic

fluids would lower the vapor pressure unless the amount of vapor is large

compared to the surface area or the volume of silicate liquid. If the

gas phase is fairly large, it must exist in voids among the solid phases,

and must support the total load pressure minus that supported by the strength

of the adjacent grains. Only if the grains have very little strength at

the depth involved, or if they are held nearly out of contact with one

another by the vapor can it be stated that vapor pressure is as great as

total rock pressure, although that relation is usually assumed. Tensile

strength within the rocks would allow vapor pressure to rise substantially

above the load pressure locally, but such excess pressure probably could not

be maintained for long if the rocks are being deformed by intrusive movements

or are being fractured tectonically.

The actual fugacity of any component in a gas mixture is a different problem, and depends upon the concentrations of the components present at any given temperature. Experimental data on fugacity coefficients in gas mixtures are scarce. However, Walter (1963) has shown that the fugacity of CO_2 in CO_2-H_2O vapors is substantially greater than that expected from its molar concentration. From considerations to be described below, however, it appears that the vapor composition in the pyroxene-hornfels facies rocks was too reducing to permit CO_2 as a major constituent.

As the 80 percent of the water produced in the reaction to hypersthene escaped and the rocks began to decrease in volume, the grains should have contacted each other irregularly and exerted some strength (sandstones commonly have some porosity at depths of 2 to 3 miles). It seems reasonable that for short periods of time the grains would have supported some of the load as they compacted. Water pressure probably fell below load pressure very rapidly as the grains started to support each other.

The originally fine-grained tuffs around the contact at Dicks Lake were deformed as they were recrystallizing during intrusion. The hypersthene-bearing rocks commonly have a monoclinic fabric. Schistosity is parallel to the contact around the intrusion from the Velma Lakes to Mt. Tallac. The hypersthene poikiloblasts in Figure 8 lie with c-axes aligned subhorizontally in the foliation plane. These c-axes define a b fabric axis that developed in response to the vertical drag of the intrusion against its walls.

Table 4. Analysis and mode of pyroxene hornfels specimen 618.

	weight percent	volume percent	
SiO_2	47.77	tan hornblende	41
Al_2O_3	13.86	plagioclase An_{60}	32
TiO_2	0.84	red-brown biotite	17
Fe_2O_3	1.96	hypersthene	$\frac{10}{100}$
FeO	9.33		
MnO	0.22		
MgO	11.81	hornblende n_Z = 1.68	
CaO	10.29	$2V_X$ = 86°	
Na_2O	1.21	2 C = 16°	
K_2O	1.23	hypersthene En_{60}	
P_2O_5	0.16		
H_2O^+	1.36		
CO_2	$\underline{0.08}$	A = 19.9	
total	100.10	C = 24.3	
		F = 55.8	

Analysis by M. Chiba, Tokyo, Japan.

Other reactions. -- A reaction from tan hornblende to hypersthene took place in an undersaturated mafic dike, specimen 618. Which other phases participated in the reaction is unknown. The texture of the hypersthene is highly poikiloblastic, as is common elsewhere in the pyroxene-hornfels facies rocks. An analysis and mode of specimen 618 is given in Table 4.

Diopsidic augite occurs with or without hypersthene in the pyroxene-hornfels facies rocks; biotite is always present. A mode and calculated analysis of a calcic andesine-diopsidic augite- red brown biotite-K feldspar-quartz hornfels is given in Table 5. The rock has more CaO and less FeO than specimen 613 (Table 3); it is the excess of CaO and not a lower temperature which caused the crystallization of augite instead of hypersthene.

Table 5. Mode and calculated analysis of pyroxene-hornfels specimen 683.

	volume percent		weight percent
quartz	2.6	SiO_2	52.8
plagioclase	48.5	Al_2O_3	18.4
K feldspar	9.0	TiO_2	1.2
biotite	24.3	Fe_2O_3	0.4
augite	15.5	FeO	5.2
total	99.9	MgO	6.1
		CaO	8.0
plagioclase An_{45}		Na_2O	3.1
augite $Ca_{41}Mg_{39}Fe_{10}$		K_2O	3.5
Density 2.840		H_2O^+	1.0
		total	99.7

$$A = 21.9$$
$$C = 28.4$$
$$F = 49.7$$

RETROGRADE EFFECTS

Within 100 to 200 feet of the contacts of the Glen Alpine
and Dicks Lake granodiorites, nearly all of the orthopyroxene and
much of the clinopyroxene has been altered to tan or green hornblende
with some magnetite. This alteration evidently was caused by late
hot solutions emanating from the plutons. The retrograde hornblende
is optically but not texturally identical with the prograde hornblende.
(Compare Figures 5 and 6.) It is the retrograde hornblende which has
the classical hornfelsic sieved texture. The small inclusions in the
large spongy hornblende grains are plagioclase and quartz, indicating
that the retrograde reaction was essentially the reverse of the prograde
reaction (1).

Farther from the contacts than about 200 feet, the pyroxenes
are fresh or are altered to pseudomorphous fibrous actinolite. No
hornblende has been altered to actinolite, however, suggesting that the
reverse of the reaction of actinolite plus an aluminous mineral to
hornblende is not easily reversible.

The water content of the retrograde hornfelses is commonly
between one and two percent. Their volume is equal to 10 percent or
less of the plutons, so that the retrograde minerals account for the
equivalent of less than 0.2% of water by weight from the magmas.

The bulk of the evidence available from the rocks of the
Mt. Tallac roof remnant indicates that most or all of the heat
responsible for the contact recrystallization was transferred by
radiation and conduction, and not by condensation of vapors driven
from the magma. The recrystallization of the highest-grade rocks
took place while magmatic movements were still proceeding, as evidenced
by the fabric of the contact schists of the pyroxene hornfels facies
(Figure 8). The Dicks Lake granodiorite has about 30 percent poikilitic
crystals (K feldspar, quartz, some hornblende and biotite) which could
not have survived post-crystallization movements, so the magma must
have been at least one-third liquid during the period of maximum
temperatures in the wallrocks. From considerations on the oxidation state
of the ferromagnesian minerals detailed elsewhere (Loomis, 1963), the
magma probably did not contain enough water to become saturated until
much more than two-thirds of the body was solid, and water would not
have been expelled until that time. Excess magmatic water in the
wallrocks would have greatly impeded the reaction to the orthopyroxenes; the
distribution in plan of the pyroxene hornfels facies strongly suggests
that the high-grade rocks were prohibited from forming until the contact
zone had become sufficiently dry.

Jaeger (1959, p. 47) pointed out that "no thermal reason exists"
for narrow contact zones around large intrusions. The fact that relatively
narrow contact zones are common may only indicate that the rate of flow
of water out of the walls controlled the rates of reactions from hydrous
to anhydrous assemblages.

The magma may have drawn in water from the rocks near the
contact, as suggested by Goranson (1937), and not expelled it until a
late stage. If the magma was not saturated when it arrived at its
present position and the water in the wall rocks was at a higher pressure
than that in the melt, water would diffuse into the magma until the
chemical potential of the water was the same on both sides of the
contact. The experimental data by Tuttle and Bowen (1958, p. 58) on
the solubility of water in granitic liquids can be evaluated for the
heat of solution of the water (Lewis and Randall, 1923, p. 229)

$$\Delta H_{soln} = \left(\frac{\partial \ln N}{\partial T}\right)_P RT^2$$

where ΔH is the differential heat of solution, N is the mol fraction
of water dissolved, T is the temperature in °K, R is the gas constant,
and P is the total pressure. The result is that the solution of water is exo-
thermic and of the order of 10^1 to 10^2 cal/mole.

The possibility, suggested by Sosman (1950), that thermal
transpiration would act to move water vapor toward the magma appears not
to be valid. Thermal transpiration is the process in which gas molecules
move from a region of lower to higher temperature when enclosed in volumes
small relative to the molecular mean free path. Loeb (1961, p. 354) pointed
out that the size and geometry of the pore spaces is critical in maintaining
large temperature gradients over the distance of a mean free path. Water
vapor at 700° C and 1400 bars has a density of about 0.4 g/cc, however; in a
fluid that dense the mean free path is similar to the molecular diameter which
is about 4×10^{-8} cm. Clearly, water molecules cannot move about in pores
smaller than their diameter, even if fractures of the order of 10^{-9} cm
should exist.

COMPOSITIONAL CONTROL OF CLINOPYROXENE

Diopside or augite occurs in the Mt. Tallac remnant in the actinolite-hornfels, hornblende-hornfels, and pyroxene hornfels facies. As shown in Figure 4, the clinopyroxene coexists with actinolite and clinozoisite in metasedimentary rocks which occur with actinolite-bearing mafic rocks. It commonly occurs without actinolite, however, in rocks which have excess calcite. The assemblages quartz-diopsidic augite-actinolite and quartz-diopsidic augite-calcite are interbedded.

Augite occurs as a product of the reaction of hornblende with anorthite and quartz in rocks of the high-grade hornblende-hornfels facies, suggesting that the control on its appearance is primarily thermal. However, the assemblage augite-plagioclase-biotite-quartz occurs immediately surrounding epidote clasts in rocks which are otherwise hornblende-plagioclase-biotite-quartz. The epidote nodules acted as sources of excess CaO, causing the crystallization of augite at a lower temperature than in the normal tuff breccias.

Augite occurs with or without hypersthene in the pyroxene hornfels facies rocks. Specimens 613 and 618 with hypersthene alone, have molecular ratios $\frac{CaO}{(Mg,Fe)O} = 0.35$ and 0.43, respectively (Tables 3, 5). Specimen 683, with augite alone, has a molecular ratio $\frac{CaO}{(Mg,Fe)O} = 0.57$ (Table 4). A narrow compositional field within which both augite and hypersthene occur exists where the molecular ratio $\frac{CaO}{(Mg,Fe)O} \approx 0.50$.

The general relations of the metamorphic minerals that occur

within the Mt. Tallac roof remnant are shown diagrammatically in

Figure 11. The diagram is applicable for rocks which are saturated or

nearly saturated with SiO_2. Carbonate and ultramafic rocks are not

included. The temperatures on the ordinate are estimates based upon the

phase equilibria given by Boyd (1959) and Greenwood (1963), the temperature

determined for the epidote-andradite reaction, and the considerations outlined

in the following section.

PHYSICAL CONDITIONS

Load pressure. — An estimate of the lithostatic pressure

which existed during the contact metamorphism can be made by two methods

which are independent of the mineral assemblages in the rocks. One is

to project rock structures above the present surface; the second is

to estimate the depth of erosion from details of the Tertiary chronology

and definition of geomorphic surfaces in the surrounding area.

No pluton in the area shows evidence of an arched or domical

structure at the exposed level; all have a discernable foliation which

parallels contacts and which is commonly steeper than 70°. Most contacts

dip under the roof remnant, however, indicating that the plutons were

closing structurally at higher levels. The structure of the Camper Flat

granodiorite requires at least 15,000 feet of rock above the present surface

if the pluton had a convex top. The dimensions of the structural cells of

the Wrights Lake granodiorite, which is slightly to the west of the area

shown in Figure 1, would also require about 15,000 feet to close into an

arched form. The rocks exposed in the core of the syncline at the top of

Mt. Tallac when projected to a point above the pyroxene hornfels facies rocks

would lie 12,000 feet above the present level.

Details of the geomorphology and Tertiary chronology of the
northern Sierra Nevada indicate a maximum depth of erosion of about
25,000 feet since intrusion of the youngest granitic rocks (Loomis, in press).
The total overburden, then, can be estimated at 20,000 to 25,000 feet,
which would correspond to about 1500 bars if the average rock density
were 2.5 g/cc.

Temperature. -- The maximum temperature reached during recrystalliza-
tion of the pyroxene-hornfels facies rocks may be estimated in several ways,
as follows.

a) The highly sporadic and patchy occurrence of K feldspar,
and the large, spongy, interstitial nature of the quartz
in many specimens suggest that these minerals had begun to
melt. At 1000 atm P_{H_2O}, melting might begin above 720°
(Tuttle and Bowen, 1958, p. 54-57).

b) The ratio of Ab in plagioclase to Ab in coexisting K feldspar
is, at any given pressure, a potentially valuable thermometer
(Barth, 1956). K feldspars in the biotite-hypersthene contact
rocks have $2V_X$ from 47° to 64°, averaging 54°. They are only
slightly perthitic near the edges of the crystals and are
untwinned. Using Tuttle's curve (1952) for a gross
approximation to the Ab content of the feldspar, and taking
the plagioclase as $An_{57}Ab_{40}Or_3$, Barth's 1956 curve gives
850°-900°; this seems too high.

c) If the contact is considered a plane in the small area west of Upper Velma Lake and Fontanillis Lake, and the intrusion as an instantaneously intruded sheet with thickness of about 1000 meters, the maximum temperature of the country rocks 200 meters from the contact would be about 75 percent of the initial contact temperature (Jaeger, 1959, p. 47). If the initial temperature was 1000° for the quartz diorite, the temperature at 200 meters would have been about 750° for Jaeger's model.

d) Hypersthene has diopsidic exsolution lamallae parallel to the optic axial plane only, indicating that it is not inverted pigeonite (Hess, 1951).

A reasonable figure for the maximum recrystallization temperature of the pyroxene-hornfels facies rocks is 750°.

The recrystallization temperature of the two hornblende hornfels facies rocks 444 and B (Table 1) can be roughly estimated. The epidote to andradite reaction (reaction 3) appears experimentally to occur between 575° and 600° at 2000 bars P_{H_2O}. That reaction is observed in the field to have begun in rocks which lie between the hornblende-bearing hornfelses and the clinopyroxene-biotite hornfelses. The spread in maximum recrystallization temperature between specimens 444 and B can be estimated as 100° to 130° by applying Jaeger's (1959) thermal models, as in (c) above. The estimated recrystallization temperatures would then be about 525° for the hornblende-bearing rock(444) and about 625° to 650° for the clinopyroxene-bearing rock(B).

Relative temperatures. -- Relative final recrystallization temperatures for the analyzed rocks can be calculated if the assumption is made that the juvenile hydrogen pressure was equal in all of the rocks during metamorphism. The ferric to ferrous iron ratio expressed in mol fractions is a function of the partial pressure of oxygen, P_{O_2}, and the temperature.

Kubaschweski and Evans (1958, p. 338) gave $\Delta F = 74,620 - 29.9 T$ for the reaction $Fe_3O_4 \rightleftharpoons 3FeO + 1/2O_2$. From this and the relation $\Delta f = RT \ln K$, K can be calculated for any temperature. The equilibrium constant $K = \dfrac{(FeO)^3 \ (O_2)^{1/2}}{(Fe_3O_4)}$. The mol fractions of Fe_3O_4 and FeO will be assumed equal to their respective activities, and the partial pressure of oxygen equal to its fugacity; both approximations are sufficiently accurate for this calculation.

The P_{O_2} that obtained during the final recrystallization of the pyroxene-hornfels facies rock (Table 3) can be calculated in the manner shown by Kennedy (1948). When the analysis is recalculated to moles, FeO = 7.70 percent, and $Fe_2O_3 = 0.11$ percent of the rock. For the temperature 750°C, $K = 10^{-9.60}$.

$$(K_{750°})^2 = \frac{(FeO)^6 \ (P_{O_2})}{(Fe_3O_4)^2} = \frac{(FeO - Fe_2O_3)^6 \ (P_{O_2})}{(Fe_2O_3 + \text{an equal amount FeO})^2} \ ,$$

and

$$P_{O_2} = \frac{(10^{-9.60})^2 \ (2.20 \times 10^{-3})^2}{(7.59 \times 10^{-2})^6} = 10^{-17.8} \ \text{atm.}$$

This very low P_{O_2} must have been in equilibrium with hydrogen and water vapor in the reaction (written as a dissociation here) $H_2O \rightleftharpoons H_2 + 1/2\ O_2$. If all the hydrogen present in the rocks came from this dissociation, the water pressure can be calculated from the relation

$$P_{H_2O} = \frac{(P_{H_2})\ (P_{O_2})^{1/2}}{K750°}$$

and the value of the equilibrium constant K. K for 1 atm total pressure can be calculated from the data given by Kubaschweski and Evans (1958, p. 339) in the same way as the above calculation for K for the oxidation of iron was made. The pressure of water calculated with this equilibrium constant must be multiplied by the correction factor of $\left[\frac{KP}{2}\right]^{2/3}$ because there are more moles on the right side of the dissociation equation than the left (increasing total pressure tends to inhibit the dissociation). The water pressure derived in this way for specimen 613 is 10^{-14} atm, an absurdly low figure. Excess hydrogen, over the above that which would come from the dissociation of water, must have been present. For the water pressure to have been 1000 atm, hydrogen must have been 6 volume percent of the total, or about 60 atm. Six percent is a higher figure than those reported in analyses of volcanic rocks.

If the 6 percent of hydrogen is assumed to have been present in all of the contact rocks at this level, relative temperatures of recrystallization can be calculated by finding the temperature at which an equilibrium constant and its derivative P_{O_2} satisfy the relation

$$K = \frac{(P_{H_2})\ (P_{O_2})^{1/2}}{(P_{H_2O})} \quad \text{at a given total pressure of } H_2O + H_2 + O_2.$$

If the temperature of the pyroxene-hornfels facies rocks was 750° at $P_{H_2O} = 1000$ atm, the temperature of recrystallization of specimen 444 (Table 1 and Figure 5) was 530°, which is close to that which was estimated from the experimental data and thermal models, as explained above.

Figure 1. Index map showing location of roof remnant.

Figure 2. Geologic map.

Figure 3. Actinolite pseudomorphs after clinopyroxene showing complex twinning. Dark patches in some grains are relict augite. Plagioclase is labradorite. Specimen 584.

Figure 4. Metamorphic diopsidic augite ($2V_Z = 50-55°$) and pale green prismatic actinolite. Specimen 519.

Figure 5. Prograde hornblende from mafic meta-andesite hornfels in Glen Alpine aureole. Specimen 444.

Figure 6. Retrograde hornblende after augite in mafic meta-andesite hornfels in Glen Alpine aureole. Specimen 440.

Figure 7. Relict subhedral plagioclase form retained in clinopyroxene-bearing hornfels, Glen Alpine aureole. Specimen B.

Figure 8. Fabric of labradorite-hypersthene-biotite-quartz contact schist. c-axes of the hypersthene porphyroblasts are aligned in plane of photograph. The c-axis lineation is subhorizontal and trends parallel to the contact.

Reaction relation shown between biotite and hypersthene; very little biotite exists inside the porphyroblasts relative to plagioclase as compared with the surrounding rock. Specimen 614.

Figure 9. Detail of hypersthene-biotite relationship. Specimen 613.

Figure 10. Hypersthene porphyroblast surrounded by K feldspar concentration. Hypersthene has exsolution lamellae of diopside parallel to the optic plane only. Specimen 614.

Figure 11. Estimated temperature-composition relationships among mafic metamorphic minerals from Mt. Tallac hornfelses.